

1 **A seismic analysis of subglacial lake D2 (Subglacial Lake Cheongsuk)**
2 **beneath David Glacier, Antarctica**

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12 **Abstract.** Subglacial lakes beneath Antarctic glaciers are pivotal in advancing our understanding of cryosphere dynamics,
13 basal hydrology, and microbial ecosystems. We investigate the internal structure and physical properties of Subglacial Lake
14 D2 (SLD2), which is located beneath David Glacier in East Antarctica, using seismic data acquired during the 2021/22 austral
15 summer. The dataset underwent a comprehensive processing workflow, including noise attenuation, velocity analysis, and
16 prestack time migration. The migrated seismic sections revealed distinct reverse-polarity reflections at the glacier–lake
17 interface; however, reflections from the lake–bed sediment interface were ambiguous, leading to interpretational uncertainty
18 about the presence of a sediment layer. To resolve this interpretational uncertainty, two alternative structural models were
19 established: Model 1 (no sediment) and Model 2 (with a sediment layer). Synthetic seismograms generated by wave-
20 propagation modeling were compared with field data to validate the subglacial lake structure. The results confirmed the water
21 column thickness to be approximately 82 m (Model 1) or approximately 10 m (Model 2), and possible structural scenarios for
22 the subglacial lake were presented. Additionally, discontinuous reflections detected in seismic sections transverse to the ice
23 flow were interpreted as scour-like feature surfaces formed by ice movement. This study identified the basal structure beneath
24 the subglacial lake, which had been challenging to identify with conventional radar surveys, through seismic surveying. In
25 addition, ambiguous signals in the field seismic data were mitigated via quantitative comparison with synthetic data, thereby
26 facilitating interpretation of the underlying structure. Collectively, these findings enhance our understanding of subglacial lake
27 environments and inform the selection of future drilling sites for in situ sampling.

28 **1 Introduction**

29 Subglacial lakes beneath the Antarctic ice sheet are typically overlain by glaciers several kilometers thick and have remained
30 isolated from direct atmospheric and solar influences for millions of years, creating extreme environments characterized by
31 low temperatures (Thoma et al., 2010) and high pressures (Tulaczyk et al., 2014). With increasing scientific interest, subglacial

32 lakes have become a focal point for studies related to the Antarctic paleoclimate, as inferred from lake sediments, as well as
33 investigations into microbial life in polar ecosystems (Bell et al., 2007, 2011; Bentley et al., 2009; Christner et al., 2014;
34 Engelhardt et al., 1990; Priscu and Christner, 2003; Rose, 1979; Wingham et al., 2006). Subglacial lakes in Antarctica are
35 generally categorized as either stable or active. Approximately 80% of subglacial lakes in Antarctica are classified as stable
36 subglacial lakes. These closed systems do not exhibit significant surface elevation changes and are characterized by long-term
37 balance between recharge and discharge, although the extent of subglacial water exchange remains uncertain in the absence of
38 direct observations. The remaining 20% are classified as active subglacial lakes, which exhibit surface elevation changes due
39 to episodic water drainage and refilling events (Livingstone et al., 2022). Such active lakes can reduce basal friction as they
40 expand, thereby facilitating glacier flow and, in some cases, accelerating calving processes, ultimately influencing glacier
41 dynamics (Bell et al., 2007; Stearns et al., 2008; Winsborrow et al., 2010). Characterizing subglacial lakes is essential for
42 understanding cryospheric processes, reconstructing past climate conditions, and assessing the potential for life in isolated,
43 extreme environments.

44 The sampling of subglacial lake water, sediments, and microbial communities is critical to address these scientific objectives.
45 However, successful sampling requires careful selection and characterization of the drilling site. Airborne ice-penetrating radar
46 (IPR) surveys are commonly employed at regional scales to detect potential subglacial lakes suitable for drilling (Christianson
47 et al., 2012; Lindzey et al., 2020; Yan et al., 2022). However, due to signal attenuation in water, IPR surveys are limited in
48 resolving the internal structure of subglacial lakes. To overcome this limitation, seismic surveys have been conducted at
49 potential subglacial lake candidates identified from IPR surveys. During such surveys, P-waves propagate through the water
50 column and are partially reflected at the lake–bed interface because of contrasts in acoustic impedance. Analyzing these
51 reflected waves enables detailed delineation of the water column and underlying substrate, thereby informing optimal drilling
52 locations (Brisbourne et al., 2023; Filina et al., 2008; Horgan et al., 2012; Woodward et al., 2010).

53 As such, numerous studies have utilized seismic surveys to investigate the characteristics of subglacial lakes, including
54 Subglacial Lake Ellsworth, Subglacial Lake Whillans, and Subglacial Lake CECs. Subglacial Lake Ellsworth, located beneath
55 2,930–3,280 m of glacial ice in West Antarctica, was the subject of a seismic survey during the austral summer of 2007–08.
56 This survey revealed spatially variable ice thickness and a lake water column ranging from 52 to 156 m, which guided the
57 identification of an optimal drilling location (Smith et al., 2018; Woodward et al., 2010). Subglacial Lake Whillans lies beneath
58 approximately 800 m of ice. Seismic observations conducted during the 2010/11 field season revealed water columns
59 extending over a 5 km segment of the survey profile, with a maximum thickness of less than 8 m. The glacier bed was
60 predominantly composed of soft sediments, and localized zones with shallow water columns (< 2 m) were also identified
61 (Horgan et al., 2012). Subsequent drilling in the summer of 2012/13 confirmed the presence of microbial life in both the water
62 and sediment samples (Christner et al., 2014). Subglacial Lake CECs (SLCECs), located beneath 2653 m of ice at the Rutford–
63 Institute–Minnesota Divide in West Antarctica, were investigated through seismic surveys conducted in the 2016/17 and
64 2021/22 seasons. These surveys revealed a maximum water column thickness of 301.3 ± 1.5 m and clastic sediments up to 15

65 m thick covering the lakebed. While the lake center was relatively flat, significant topographic variability was observed near
66 the lake margins (Brisbourne et al., 2023).

67 We have initiated subglacial lake research beneath David Glacier, the closest major glacier to Jang Bogo Station in East
68 Antarctica. Satellite altimetry has identified six subglacial lakes in this region (Smith et al., 2009; Wright and Siegert, 2012).
69 During the 2016/17 austral summer, an airborne IPR survey was conducted over the region encompassing Subglacial Lake D1
70 (SLD1) and Subglacial Lake D2 (SLD2) (Lindzey et al., 2020). A subsequent high-resolution IPR survey was carried out
71 during the 2018/19 field season, focusing solely on SLD2 (also referred to as “Subglacial Lake Cheongsuk”) (Ju et al., 2025).
72 Ju et al. (2025) subdivided the previously identified single subglacial water body at SLD2, as detected by ICESat altimetry,
73 into three smaller subglacial lakes: SLD2-A, SLD2-B, and SLD2-C. Among these, SLD2-A represents the largest areal extent,
74 and targeted seismic surveys were conducted over this area to obtain high-resolution information on the lake depth and basal
75 structure. In the 2019/20 season, an initial seismic campaign identified the glacier thickness and suggested the presence of the
76 lake; however, the data quality was compromised by surface crevasse noise and a lack of adequate fold coverage, limiting
77 detailed interpretation. A refined seismic survey with 8-fold coverage was conducted during the 2021/22 season to address
78 these issues. Furthermore, the sound source was positioned further from the crevasse (end-shot), delaying the arrival of
79 crevasse-generated noise and preventing it from obscuring key reflections.

80 In this study, we present a detailed analysis of the physical and structural properties of SLD2-A using seismic data acquired
81 during the 2021/22 campaign. We first describe the seismic data processing workflow, including noise attenuation, amplitude
82 correction, and prestack time migration. Some areas of the processed field seismic data are challenging to interpret due to a
83 lack of subsurface information, overlap with ghost signals, and signal attenuation. In the case of the SLD2 region, the absence
84 of borehole data introduces inherent uncertainty into the subglacial lake structure derived from the Prestack Time Migration
85 (PSTM) section. In particular, reflections associated with the sediment layer are challenging to interpret because they have
86 weak amplitudes and overlap with ghost components. To compensate for these limitations, a subsurface structural model was
87 constructed, and model-based synthetic seismograms were compared and analyzed against field observations. As a result, the
88 substructure of SLD2-A is quantitatively presented as two possible scenarios: Glacier–Lake–Bedrock (model 1) or Glacier–
89 Lake–Sediment–Bedrock (model 2).

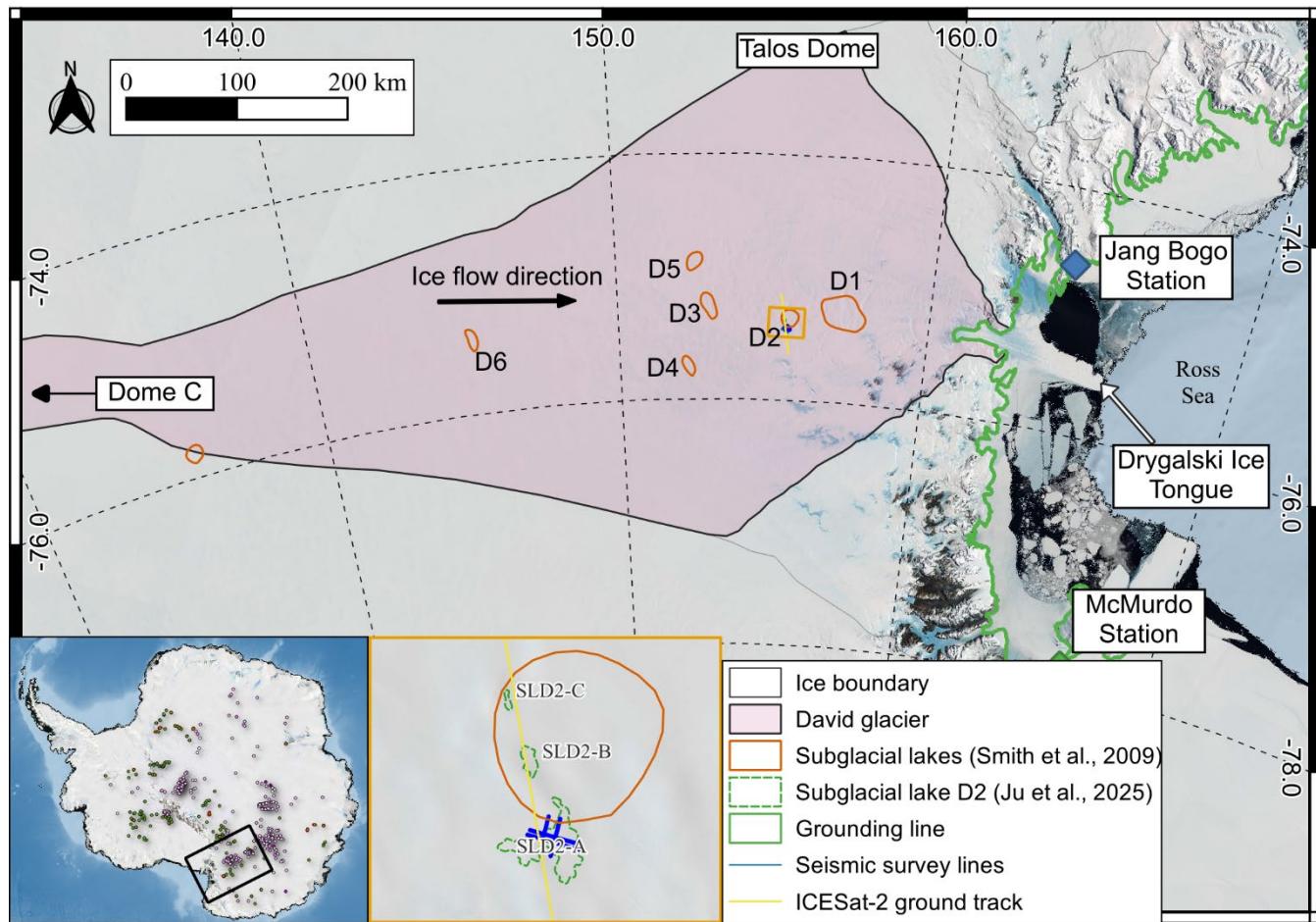
90 **2 Subglacial Lake D2 Beneath David Glacier in Antarctica**

91 **2.1 David Glacier**

92 David Glacier, located in Victoria Land, East Antarctica, originates from the Dome C and Talos Dome regions and flows
93 seaward through the Drygalski Ice Tongue (Fig. 1). The mass balance of glaciers from 1979 to 2008 has been estimated at $7.5 \pm 0.4 \text{ Gt yr}^{-1}$ (Rignot et al., 2019), while the mean ice discharge over the more extended period from 1979 to 2017 was reported
94 to be approximately 9.7 Gt yr^{-1} (Frezzotti et al., 2000; Rignot et al., 2019). According to Smith et al. (2020), satellite altimetry
95 observations from ICESat-1 and ICESat-2 (2003–2019) indicate that the grounded portion of David Glacier experienced a mass
96

97 gain of 3 ± 2 Gt yr $^{-1}$, whereas the adjacent ice shelves exhibited a mass loss of -1.6 ± 1 Gt yr $^{-1}$. Although the overall mass
98 balance of David Glacier currently appears stable, several active subglacial lakes observed by satellites have the potential to
99 influence glacier dynamics (Ju et al., 2025; Kim et al., 2025).

100



101
102 **Figure 1: Locations of subglacial lakes D1–D6 in the David Glacier region, Victoria Land, Antarctica (EPSG: 4326–WGS84).**
103

104 **2.2 Subglacial Lake D2**

105 Among the six subglacial lakes (D1–D6) identified beneath David Glacier via satellite altimetry (Smith et al., 2009; Wright
106 and Siegert, 2012), SLD2 was observed to have experienced a drainage event between 2003 and 2008 on the basis of ICESat
107 altimetry data (Smith et al., 2009). Since the drainage event, a continuous increase in surface elevation over SLD2 has been
108 observed, indicating water refilling, as detected from CryoSat-2 altimetry data (2013–2017) (Siegfried and Fricker, 2018) and,
109 more recently, from ICESat-2 observations (2019–2024) (Fig. 2). Figure 2 shows elevation changes relative to April 2019,

110 indicating surface uplift through January 2022. After this period, the surface elevation remained stable in the region originally
111 delineated as SLD2 by Smith et al. (2009), whereas a decreasing elevation trend was observed in the SLD2-A region (Ju et al.,
112 2025). These patterns of elevation change strongly suggest that SLD2 is an active subglacial lake, with cyclic drainage and
113 refilling likely contributing to the presence of subglacial sediments (Siegfried et al., 2023).

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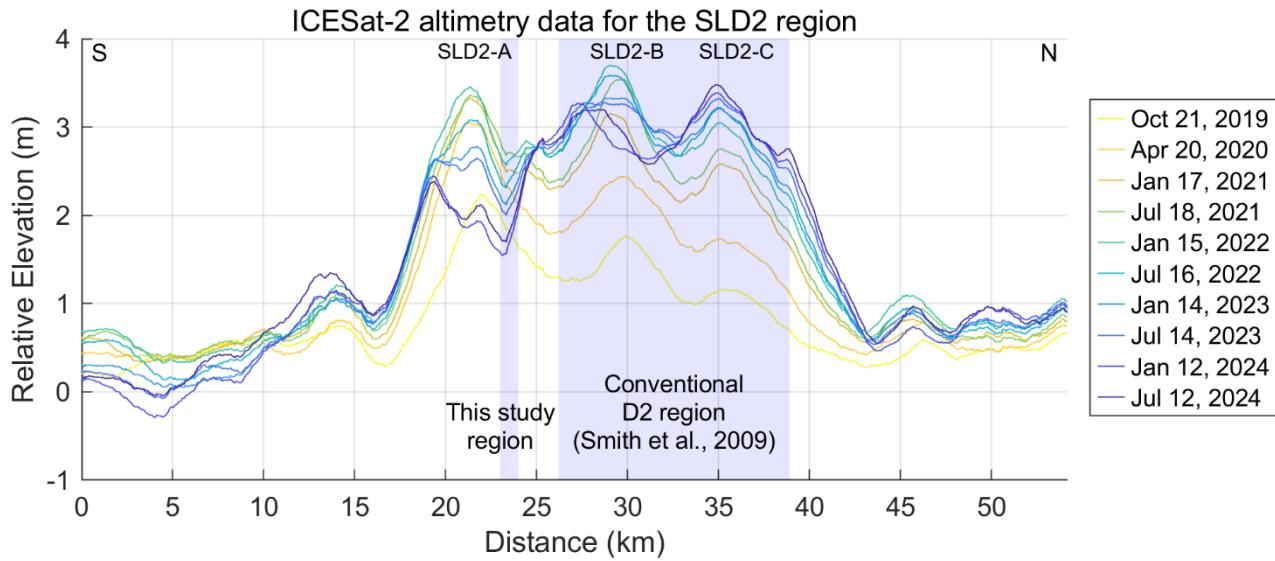


Figure 2: Glacier surface elevation changes derived from ICESat-2 altimetry between 22 April 2019 and 12 July 2024. The X-axis corresponds to the 22 April 2019 dataset, and all subsequent elevation changes are referenced to this date. The light blue shaded region indicates the spatial overlap between the conventional SLD2 region identified by Smith et al. (2009) and our study region.

119

120 To better constrain the extent and basal conditions of SLD2, we used airborne IPR data collected during the 2016/17 (Lindzey
121 et al., 2020) and 2018/19 (Ju et al., 2025) field campaigns. These surveys show that the glacier surface elevations in the SLD2
122 region range from approximately 1820 to 1940 m. The corresponding ice thicknesses vary between 1685 and 2293 m.
123 Furthermore, the observations of moderately enhanced radar bed echoes relative to the surrounding area, elevated specularity
124 values (>0.4), depressed basal elevations (≤ -350 m), the presence of a basin-like topography, a lower hydraulic head than the
125 surroundings, and low hydraulic gradients ($\leq 0.84^\circ$) collectively suggest a high potential for the presence of subglacial water
126 beneath SLD2. (Ju et al., 2025; Lindzey et al., 2020).

127 **3 Method**128 **3.1 Seismic survey**

129 As previously noted, the internal structure and water column of subglacial lakes cannot be fully resolved using IPR alone
 130 because of signal attenuation in water. Accordingly, a seismic survey was conducted within the candidate SLD2-A region
 131 identified from IPR data to investigate the structure of the subglacial lake more precisely.

132 During the 2019/20 austral summer, a preliminary seismic survey was conducted over the SLD2-A region to evaluate the
 133 potential presence of a subglacial lake and to obtain initial information on its structural characteristics. Owing to limited field
 134 time and equipment constraints, the fold of coverage for all survey lines was restricted to 1, and all shot points happened to be
 135 aligned near surface crevasses. Consequently, the acquired seismic data were significantly degraded by strong linear coherent
 136 noise generated by crevasses, severely compromising the quality of key reflectors, particularly those at the subglacial lake–
 137 bedrock interface. Furthermore, explosives were deployed in shallow boreholes (< 20 m depth), and due to the absence of
 138 proper backfilling, poor coupling between the explosives and the borehole walls further reduced energy transmission efficiency,
 139 resulting in overall low-quality reflection signals (Ju et al., 2024). Combined with the limitations of single-fold acquisition,
 140 stacking was not feasible, the dataset exhibited a low signal-to-noise ratio (SNR) and was unsuitable for quantitative structural
 141 interpretation. Nevertheless, the preliminary survey qualitatively confirmed the glacier thickness beneath SLD2-A and
 142 suggested the presence of subglacial water, providing critical baseline information that guided the methodology and survey
 143 design of the subsequent detailed seismic campaign conducted during the 2021/22 season.

144 For the refined survey, seismic acquisition lines were planned using bed topography derived from the IPR and surface elevation
 145 data from satellite altimetry. A total of four seismic lines were acquired and designated 21X, 21Y, 21XX, and 21YY (Fig. 3).
 146 Lines 21X and 21XX, oriented approximately 60° relative to the ice flow direction, are situated at an average surface elevation
 147 of 1894 ± 13 m. Lines 21Y and 21YY, oriented approximately -30° in the ice flow direction, lie at an average elevation of
 148 1887 ± 16 m. All lines traverse regions of minimal topographic relief, with average surface slopes of approximately 0.5°,
 149 indicating a relatively flat and stable glacier surface. The lengths of the 21X/21XX and 21Y/21YY lines are approximately 5
 150 km and 3.5 km, respectively. Seismic acquisition for lines 21X and 21Y was conducted using 8-fold coverage to increase the
 151 resolution, whereas lines 21XX and 21YY were acquired with 4-fold coverage due to time constraints during the survey. The
 152 additional acquisition parameters are summarized in Table 1.

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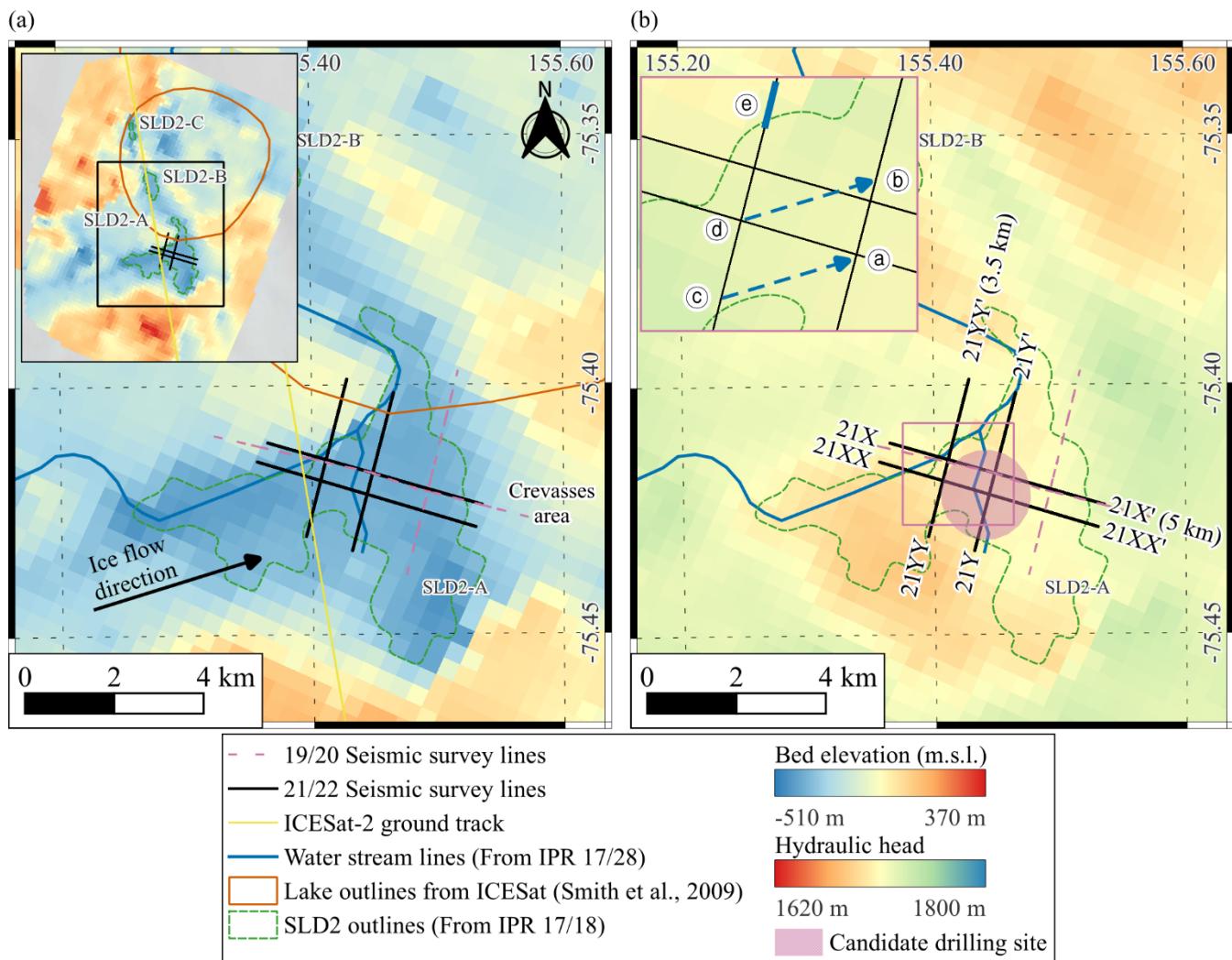
154 **Table 1: Parameters of the active-source seismic survey.**

Survey Parameters	Survey lines			
	21X line	21Y line	21XX line	21YY line
Line length (km)	5	3.5	5	3.5
Fold	8	8	4	4
Shot interval (m)	90	90	180	180
Number of shots	56	40	28	20

Shot positioning	Use both off-end and center shots
Receiver channels	96
Receiver interval (m)	15
Near offset (m)	0
Far offset (m)	1425
Recording time (s)	4
Record peak frequency (kHz)	1
Record sampling rate (ms)	0.25
Survey time (days)	34
Survey crew size	

Hot water drilling (3), Seismic (6)

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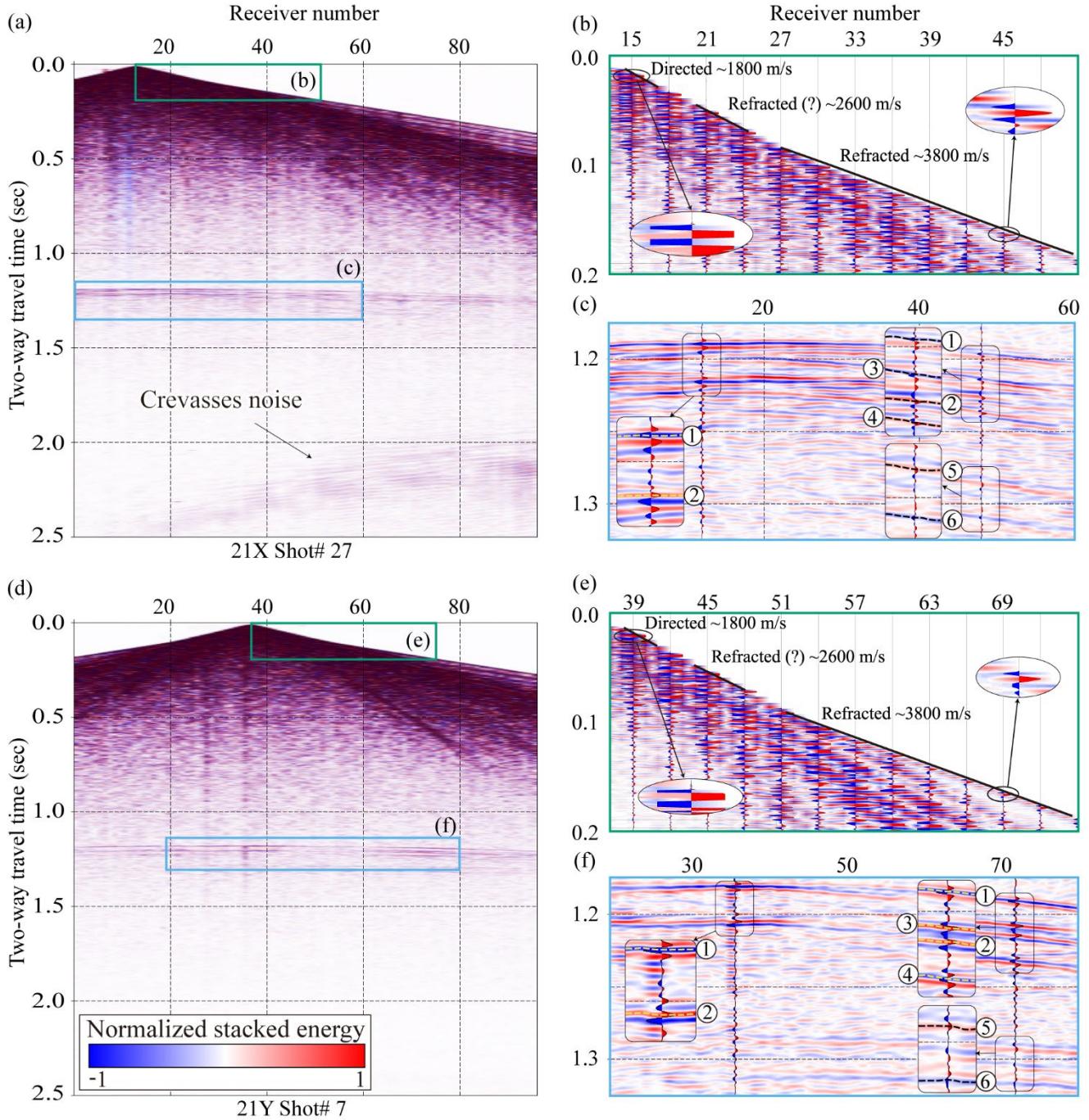
160 Before the seismic survey, a ground-penetrating radar (GPR) survey was used to identify the firn transition zone at depths of
 161 approximately 20–22 m. To enhance seismic signal transmission, 1.6 kg of pentaerythritol tetranitrate (PETN) explosives were
 162 emplaced at depths of 25–30 m using hot water drilling techniques. A total of 144 shots were deployed across the four survey
 163 lines. Detailed shot positioning information is provided in the supplementary information S1. Given the snow-covered glacier
 164 surface, Georods were used instead of conventional spike-type geophones to increase signal detection efficiency (Voigt et al.,
 165 2013). Each Georod houses four geophone elements in a 0.6 m-long cylindrical array, producing a single output by summing
 166 the inputs from all the elements. Compared with traditional geophones, this configuration improves coupling and detection
 167 performance in snow-dominated environments (Voigt et al., 2013). Figure 4 presents shot gather #27 from line 21X and shot
 168 gather #7 from line 21Y. In these shot gathers, the velocity of the direct wave is estimated to be approximately 1800 m/s, and
 169 the refracted wave velocity in firn-ice transition is approximately 3800 m/s. First-arrival analysis of the direct wave indicates
 170 a normal polarity, confirming the source waveform polarity. A prominent negative polarity reflection is observed at a two-way
 171 travel time (TWT) of approximately 1.2 s, interpreted as the glacier–lake interface (①; See Table 2 for symbols definitions).
 172 Approximately 25–30 ms later, a ghost reflection (②) with normal polarity appears. A subsequent reflection at approximately
 173 1.3 s TWT, showing normal polarity, is attributed to the bed interface (⑤), followed by its negative polarity ghost reflection
 174 (⑥) 25–30 ms later. In some shot gathers, reflection signal (③) and its corresponding ghost signal (⑤) are observed. Notably,
 175 while signal ③ generally appears with normal polarity in most records, it appears with reverse polarity in a few cases, such
 176 as Shot #27 on line 21X. The survey was designed to place the seismic source at a distance from crevasses, ensuring that
 177 crevasse-related noise would be recorded after the main reflections (1.1–1.3 s), thereby minimizing its impact (Figure 4a).
 178 While most data exhibit crevasse noise occurring after the main reflections, a reduction in the source–crevasse distance causes
 179 this noise to increasingly overlap with the primary arrivals, thereby complicating interpretation.

180

181 **Table 2: Symbols for each reflection event**

Interface symbols	Model 1	Model 2
①	Ice-water	Ice-water
②	Ice-water ghost	Ice-water ghost
③	-	Water-sediment
④	-	Water-sediment ghost
⑤	Water-bed	Sediment-bed
⑥	Water-bed ghost	Sediment-bed ghost
⑦	Ice-bed	Ice-sediment
⑧	Ice-bed ghost	Ice-sediment ghost

182



183

184 **Figure 4:** Raw shot records from seismic lines 21X (a) and 21Y (d). Panels (b) and (e) are zoomed-in views of the early arrival window
 185 (0.0–0.2 s) from panels (a) and (d), respectively, used to calculate the apparent velocities of the direct and refracted waves. These
 186 panels highlight that the first arrivals of both the direct wave (clipped for display) and the refracted wave exhibit positive polarity.
 187 The direct wave, propagating through the upper firn layer (0–25 m depth), shows an apparent velocity of approximately 1800 m/s, while the
 188 refracted wave in firn–ice transition has an apparent velocity of approximately 3800 m/s. Panels (c) and (f) are zoomed-in

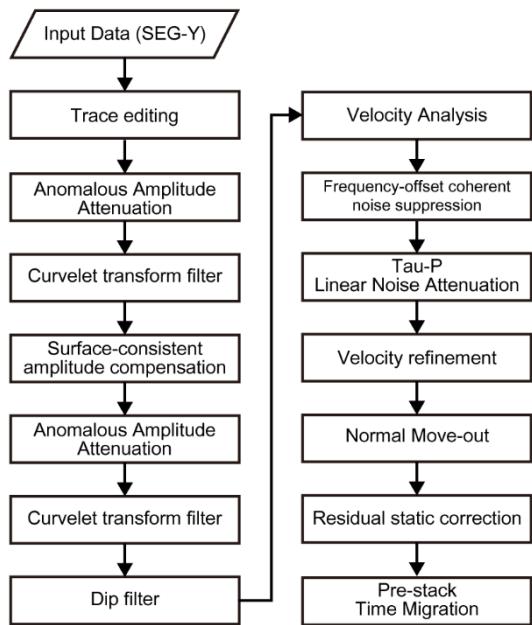
189 views of the deeper arrivals (1.1–1.4 s) from panels (a) and (d), respectively. Reflections from the ice–water (①) interface exhibit
190 negative polarity, whereas those from the water/sediment–bed (⑤) interface display positive polarity.

191

192 **3.2 Seismic data processing**

193 Although seismic data acquired from glaciers share processing similarities with those of land-based surveys, glaciological
194 factors, such as surface cracks, crevasses, and strong winds, introduce substantial noise that can degrade data quality (Johansen
195 et al., 2011; Zechmann et al., 2018). Among these factors, linear noise generated by crevasses is particularly detrimental, often
196 obscuring key reflections (Dow et al., 2013). Hence, the glacier seismic data underwent multiple data processing sequences
197 focused on linear noise removal (Fig. 5). Acquisition geometry was added to the data using the raw data and geometry
198 information. Multiple data processing and noise removal processes were then carried out to increase the signal-to-noise ratio
199 (SNR).

200



201
202 **Figure 5: Schematic of the seismic data processing workflow based on the Omega geophysical data processing platform (SLB),**
203 **including noise attenuation, amplitude correction, velocity analysis, and prestack time migration.**

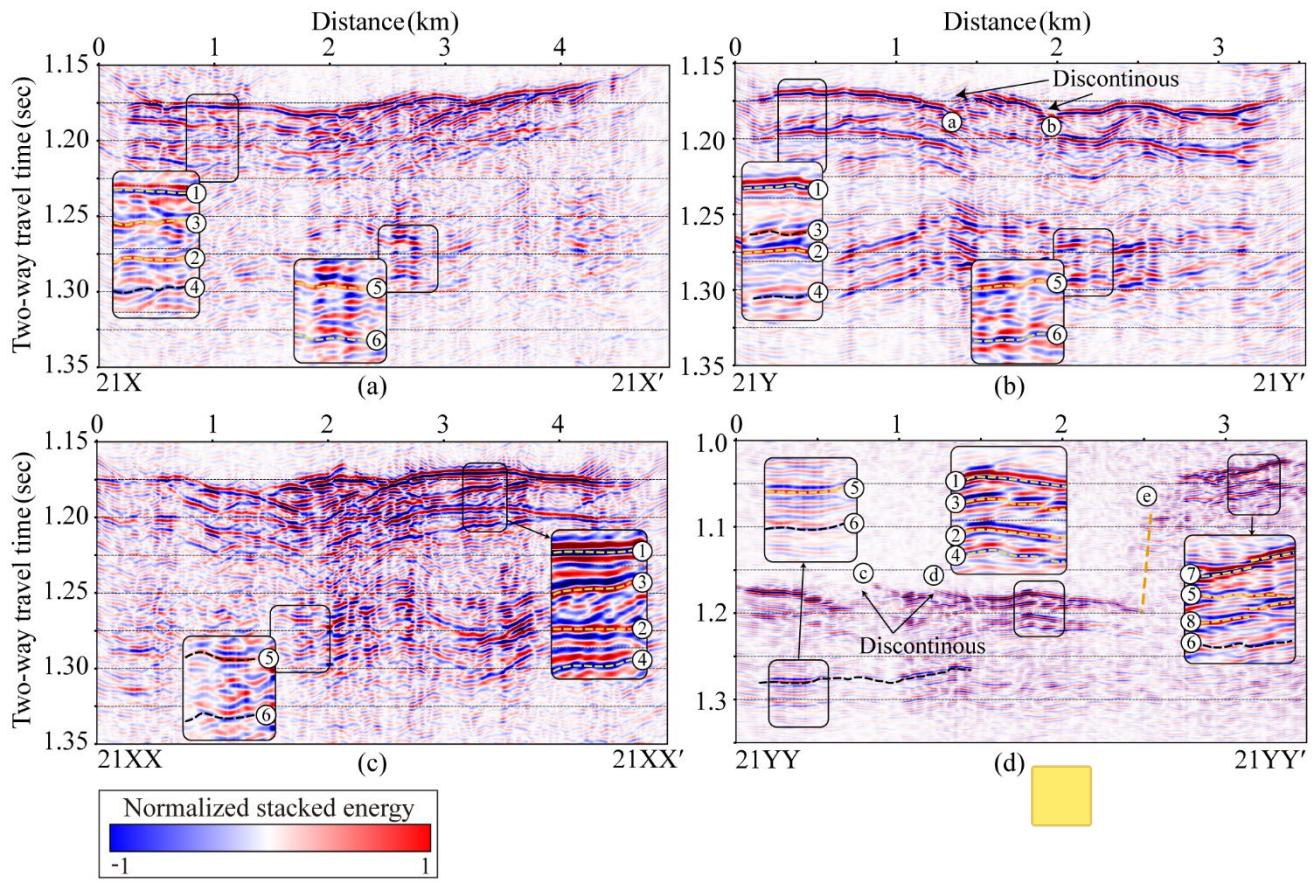
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205 The initial processing involved anomalous amplitude attenuation (AAA), implemented via a spatial median filter. This step
206 targets outlier amplitudes within a defined frequency band, attenuating anomalous signals through interpolation across
207 neighboring traces. A curvelet transform-based filter was subsequently applied to remove coherent noise. Curvelet
208 decomposition enables the separation of signals on the basis of dip angle and scale, allowing for the selective removal of

209 ground roll and other coherent noise components that differ in dip from true reflections (Oliveira et al., 2012). In this study,
210 linear coherent noise at later arrival times (>2.0 s) was effectively removed using this method.
211 Surface-consistent amplitude compensation (SCAR) and surface-consistent deconvolution were employed to normalize the
212 amplitude variability across shot gathers. These steps were followed by a second round of AAA and curvelet filtering to
213 suppress artifacts introduced during the compensation and deconvolution stages. Dip filtering was also applied to eliminate
214 spurious hyperbolic arrivals, which were manually identified and removed.
215 Velocity analysis was conducted at intervals of 40 common midpoints to construct a migration velocity model. Frequency–
216 offset coherent noise suppression (FXCNS) was used to attenuate linear-related noise, followed by Tau-p linear noise
217 attenuation (LNA), effectively reducing the noise associated with crevasse scattering. The final processing steps included
218 velocity model refinement, normal move-out (NMO) correction, and prestack time migration (PSTM). The specific parameters
219 employed for data processing, as well as the intermediate outcomes at each processing stage, are provided in the supplementary
220 information S2.
221 To increase imaging accuracy, a residual static correction was applied before migration using glacier surface elevation data.
222 The final migrated seismic section was produced using Kirchhoff PSTM. The migrated data have a center frequency of
223 approximately 180 Hz. Assuming seismic wave velocities between 1395 m/s and 3800 m/s, the corresponding vertical
224 resolutions, which are calculated using the quarter-wavelength criterion, range from approximately 2.01 m to 5.27 m. The data
225 can image both the top and bottom of a water column approximately 2 m thick or thicker.

226 **4 Seismic data processing results**

227 Figure 6 presents the PSTM results for the four seismic survey lines. On line 21X (Fig. 6a), a strong, laterally continuous
228 reflection with reverse polarity is observed at 0.3–4.8 km along the profile, and the two-way travel time (TWT) is
229 approximately 1.15–1.18 s. This reflection is interpreted as the glacier–lake interface (①). Approximately 25–30 ms below
230 this horizon, a normal polarity reflection (②) appears, likely representing a ghost signal associated with the primary glacier–
231 lake reflection. Between reflections ① and ②, a weak normal polarity reflection (③), presumed to represent an interface, is
232 observed. However, in some shot gathers, signal ③ appears with reverse polarity (Figure 4c), leading to partial cancellation
233 and ambiguity in layer interpretation. Approximately 25 ms later, an opposite polarity ghost reflection (④) follows. A deeper
234 normal polarity reflection is observed within 1.9–3.1 km at TWTs of 1.25–1.27 s (⑤), which is interpreted as the bed interface.
235 This is followed by a reverse polarity reflection 25–30 ms later (⑥), which is presumed to be the corresponding ghost of the
236 bed interface.



237
 238 **Figure 6: PSTM seismic sections for lines (a) 21X, (b) 21Y, (c) 21XX, and (d) 21YY prior to ghost removal. Ghost reflections appear**
 239 **25–30 ms beneath the glacier–lake and lake–bed interfaces due to the 25 m source depth. See Table 2 for symbols definitions.**

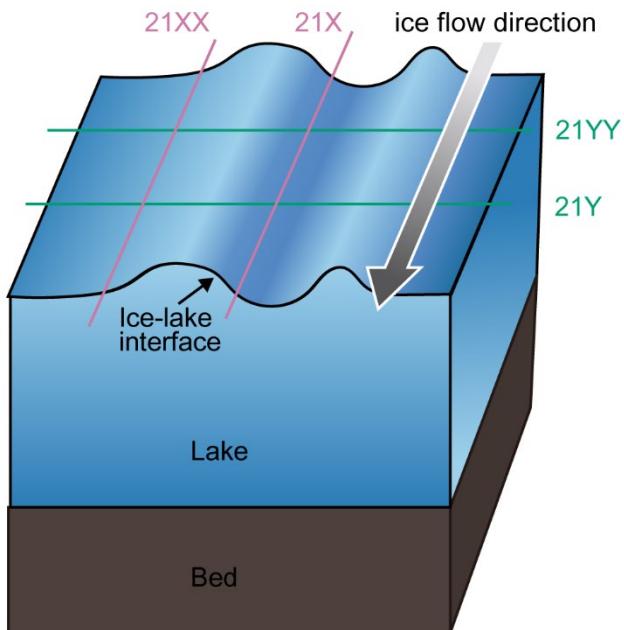
240
 241 In line 21Y (Fig. 6b), similar features are observed. A reverse polarity reflection, interpreted as the glacier–lake interface (①),
 242 is observed within 0.1–3.2 km at TWT 1.17–1.18 s, with its ghost reflection (②), which exhibits normal polarity and appears
 243 25–30 ms later. Between reflections ① and ②, a weak normal polarity reflection (③), presumed to represent an interface, is
 244 observed in some areas, followed approximately 25 ms later by an opposite polarity ghost reflection (④). A normal polarity
 245 reflection within 0.1–3.2 km at a TWT of 1.26–1.27 s is interpreted as the bed interface (⑤), followed by a reverse polarity
 246 ghost signal (⑥). Additionally, discontinuous reflections interpreted as subglacial scour-like features (SLF) are visible at
 247 approximately 1.3 km (ⓐ) and 1.9 km (ⓑ) along line 21Y at TWT 1.18 s (black arrows in Fig. 6b). These features may be
 248 associated with glacial erosion of the underlying substrate.
 249 In line 21XX (Fig. 6c), a reverse polarity reflection, interpreted as the glacier–lake interface (①), is observed within 0–4.3
 250 km at a TWT of 1.17–1.18 s. This reflection is followed 25–30 ms later by a normal polarity reflection (②), which is
 251 considered the ghost of the primary glacier–lake interface. Between reflections ① and ②, a weak normal polarity reflection

252 (③), presumed to represent an interface, is observed in some areas, followed approximately 25 ms later by an opposite polarity
253 ghost reflection (④). Further down the section, a normal polarity reflection (⑤) within 1.9–4.2 km at a TWT of 1.25–1.28 s
254 is interpreted as the bed interface, followed by its ghost reflection (⑥) 25–30 ms later.

255 On line 21YY (Fig. 6d), the glacier–lake interface (①) is marked by a strong, flat, reverse polarity reflection at 0–2.4 km and
256 a TWT of 1.17–1.20 s, followed by its normal polarity ghost (②) 25–30 ms below. A weak normal polarity reflection (③),
257 presumed to represent an interface, is observed between ① and ②, followed approximately 25 ms later by a opposite polarity
258 ghost reflection (④). Bed interface reflections (⑤) are observed within 0.2–2.4 km at TWTs of 1.27–1.29 s, followed by a
259 reverse polarity ghost (⑥) 25–30 ms later. Within 2.4–2.55 km and TWTs of 1.08–1.17 s, no coherent reflection is visible
260 due to the steeply dipping bed topography, as indicated by the dashed orange (⑧) line in Fig. 6d. Within 2.55–3.4 km and a
261 TWT of 1.03–1.09 s, a reverse polarity reflection (⑦), likely originating from a mildly dipping sedimentary surface, is
262 observed, followed by an opposite polarity ghost reflection (⑧). Additionally, although weak, reflection signal (⑤) and its
263 corresponding ghost (⑥) are also identified. Additionally, similar to observations on line 21Y, discontinuous reflections
264 interpreted as SLF surfaces appear at 0.7 km (⑨) and 1.2 km (⑩) along line 21YY at TWT 1.18 s (black arrows in Fig. 6d).

265 The discontinuous reflection signals identified on lines 21Y and 21YY are spatially aligned along the ice flow direction when
266 projected laterally (Fig. 3, dashed blue arrow). This alignment suggests that the observed discontinuities correspond to a
267 subglacial SLF surface formed by glacial motion. The SLF is visible predominantly on lines 21Y and 21YY, which are oriented
268 more perpendicularly to the ice flow direction, thereby enhancing the expression of lateral subglacial variability. In contrast,
269 lines 21X and 21XX are more parallel to the ice flow, resulting in a foreshortened view of the subglacial structures and a
270 relatively flat appearance in the seismic sections (Fig. 7).

271



272

273 **Figure 7: Conceptual diagram illustrating the orientation of seismic survey lines relative to subglacial structures and the ice flow**
 274 **direction, explaining the appearance of structural features in each line.**

275

276 **5 Comparison between field data and synthetic seismograms**

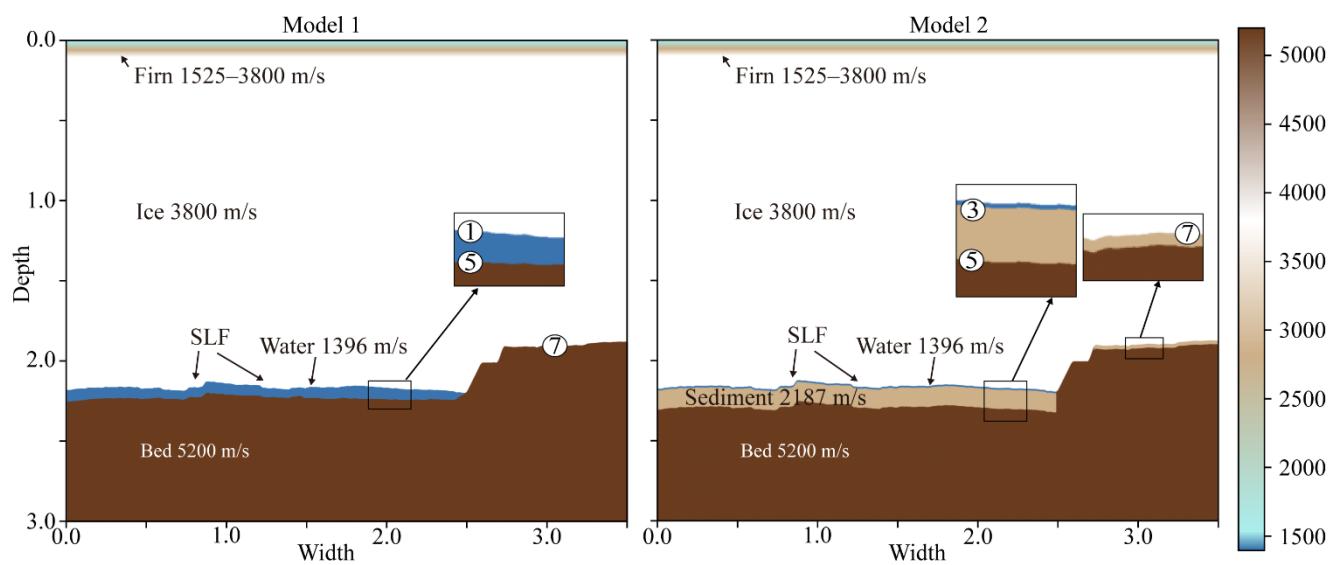
277 In all seismic profiles, the glacier–water interface (①) is characterized by strong, reverse polarity reflections. Following this,
 278 a relatively weaker reflection (③) with limited lateral continuity, which may indicate an unconsolidated sediment layer, or an
 279 unknown interface beyond the scope of current interpretation.

280 Interpreting field seismic data presents inherent challenges due to limited subsurface information, high levels of ambient noise,
 281 and signal attenuation. These issues are particularly pronounced at the SLD2 site, where the absence of borehole data
 282 introduces significant uncertainty and potential inaccuracies in depth estimations derived from PSTM sections. Such
 283 limitations may lead to misinterpretations of stratigraphic boundaries (Herron, 2000; Yilmaz, 2001). To address these
 284 challenges, this study developed a subsurface structural model and conducted a comparative analysis of synthetic seismograms
 285 generated from the model with observed field data. Focusing on the interpretation of basal reflections beneath the subglacial
 286 lake—excluding the glacier–lake interface (①)—two plausible structural models were proposed. Model 1 assumes the absence
 287 of a sedimentary layer, in which reflection (③) is not present, and reflection (⑤) represents the base of the subglacial lake.
 288 In contrast, Model 2 includes a sedimentary layer, where reflection (③) corresponds to the lake–sediment interface and
 289 reflection (⑤) indicates the sediment–bedrock interface (Figure 8). The synthetic data were generated using a time-domain
 290 forward modeling approach based on the staggered grid finite difference method (Graves et al., 1996). The velocity model

used in the simulation was constructed based on field velocity analysis and previously published data, and included stratigraphic units representing firn, glacial ice, subglacial water, sediment, and bedrock. Each layer was assigned appropriate P-wave velocities and density values. P-wave velocities in firn vary from 1525 to 3800 m s⁻¹ because density increases with depth (Kirchner and Bentley, 1979; Picotti et al., 2015; Qin et al., 2024). Glacial ice has an average P-wave velocity of approximately 3800 ± 5 m s⁻¹ at -2 ± 2 °C (Kohnen, 1974), while subglacial water has a velocity of approximately 1396 ± 2 m s⁻¹ at -1.75 ± 0.25 °C, with a salinity less than 1 PSU (practical salinity units) (Thoma et al., 2010; Tulaczyk et al., 2014). The P-wave velocities of the sediment and bed were referenced from the Lake Vostok model value (Carcione & Gei, 2003). Forward modeling was then conducted using the Ricker wavelet, with acquisition parameters matching those used in the field survey (Table 3). We applied just the migration step in case of the synthetic dataset, as it is free of noise.

301 **Table 3: Parameters of the synthetic model.**

Synthetic modeling parameters																											
Model size	3.5 km (distance) x 3 km (depth)																										
Source	Ricker wavelet (zero-phase), 60 Hz																										
Receiver	25 m depth, 90-m interval																										
Grid spacing	0 m depth, 15-m interval, 96 channel																										
Sampling interval	0.5-m																										
Layer parameters	<table> <thead> <tr> <th></th><th>Thickness (m)</th><th>Velocity (m/s)</th><th>Density (g/cm³)</th></tr> </thead> <tbody> <tr> <td>Firm</td><td>100</td><td>1525–3800</td><td>0.3–0.917</td></tr> <tr> <td>Ice</td><td>1887–2221</td><td>3800</td><td>0.917</td></tr> <tr> <td>Water (Model 1, 2)</td><td>53–82 / 10</td><td>1396</td><td>1.017</td></tr> <tr> <td>Sediment (Model 2)</td><td>120</td><td>2817</td><td>2.128</td></tr> <tr> <td>Bed</td><td></td><td>5200</td><td>3.2</td></tr> </tbody> </table>				Thickness (m)	Velocity (m/s)	Density (g/cm ³)	Firm	100	1525–3800	0.3–0.917	Ice	1887–2221	3800	0.917	Water (Model 1, 2)	53–82 / 10	1396	1.017	Sediment (Model 2)	120	2817	2.128	Bed		5200	3.2
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Bed		5200	3.2																								



304 **Figure 8: P-wave velocity model used in forward modeling for line 21YY. The upper ~100 m represents firn with velocities ranging**
305 **from 1525–3800 m s⁻¹ (Kirchner and Bentley, 1979; Picotti et al., 2015; Qin et al., 2024). The ice below this depth has a velocity of**
306 **3800 ± 5 m s⁻¹ (Kohnen, 1974), and the subglacial water layer has a velocity of 1396 ± 2 m s⁻¹ (Thoma et al., 2010; Tulaczyk et al.,**
307 **2014). In Model 2, the velocity of 2817 m s⁻¹ for the sediment layer was taken from the lower sediment layer model of Lake Vostock**
308 **(Carcione & Gei, 2003).**

309

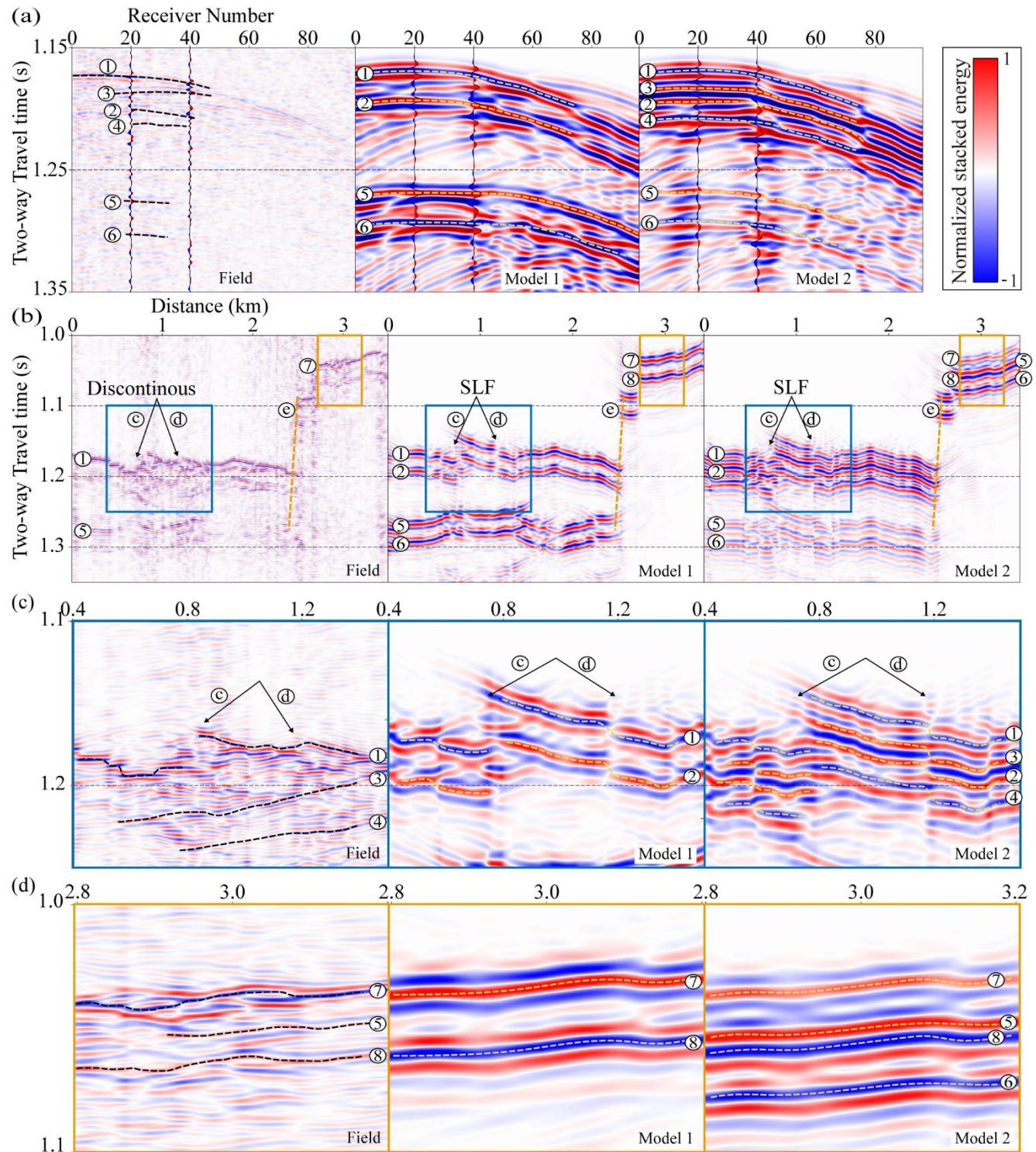
310 Figure 9a compares the shot gather from seismic data line 21YY (left) with those from the synthetic datasets for Models 1 and
311 2 (center, right). A prominent reflection at a TWT of 1.17 s is observed in both datasets, corresponding to the glacier–lake
312 interface (①). This reflection results in a high impedance contrast and reverse polarity due to the P-wave velocity difference
313 between glacial ice and water. These features are consistent with previous observations at glacier–lake interfaces (Atre and
314 Bentley, 1993; Brisbourne et al., 2023; Horgan et al., 2012; King et al., 2004; Peters et al., 2007; Woodward et al., 2010). A
315 secondary reflection with normal polarity appears approximately 28 ms after the primary event (②) and is interpreted as a
316 surface ghost reflection. This time delay corresponds to a seismic source depth of approximately 25 m, which is consistent
317 with previous seismic analyses (Brisbourne et al., 2023; Schlegel et al., 2024). That is, assuming an average P-wave velocity
318 of 1800 m s⁻¹ within the top 25 m, the TWT of the ghost reflection matches the expected delay:

319
$$\text{TWT}_{\text{ghost}} = \frac{2 \times 25 \text{ m}}{1800 \text{ m/s}} \approx 28 \text{ ms.} \quad (1)$$

320 Furthermore, considering that the acoustic impedance of air is approximately zero ($Z_{\text{air}} \approx 0$) and that of ice is Z_{ice} , the
321 reflection coefficient (RC) for an upgoing wave at the air–ice interface can be approximated as follows:

322
$$RC = \frac{Z_{\text{air}} - Z_{\text{ice}}}{Z_{\text{ice}} + Z_{\text{air}}} \approx -1. \quad (2)$$

323 This implies that the polarity of the ghost reflection at the surface is reversed relative to the downgoing primary wave (Krai
324 and Shin, 1990; Robinson and Treitel, 2008).



325 **Figure 9: Comparison of field seismic data and synthetic results from Model 1 and 2. (a) Shot gathers at the same location from the**
 326 **21YY field data (left) and synthetic models (center: Model 1, right: Model 2). (b) Comparison of PSTM images from the 21YY line**
 327

328 and the two synthetic models. (c) Enlarged views of discontinuous reflections. (d) Comparison of dipping bed reflections, showing
329 shadow zones and steep basal topography.

330

331 Figure 9b compares the PSTM sections of the field data from line 21YY and the two synthetic models. Unlike the field data,
332 the synthetic dataset is free from ambient noise and features a precise source–receiver geometry, resulting in clearer delineation
333 of subsurface reflections and facilitating structural interpretation. In the PSTM sections of both the synthetic models (Model 1
334 and Model 2) and the field data, three primary reflection events (①, ⑤, ⑦) and their corresponding source-generated ghost
335 reflections (②, ⑥, ⑧) are observed at similar two-way travel times. Reflections ①, ②, ⑤, and ⑥ also exhibit consistent
336 polarity across the synthetic and field datasets. Additionally, the lateral discontinuities in reflections generated by the SLF
337 structure implemented in the velocity model closely resemble those observed in the field data. The orange dashed line (⑩)
338 delineates the shape of the bedrock forming the margin of the subglacial lake, interpreted to dip at approximately 52°.

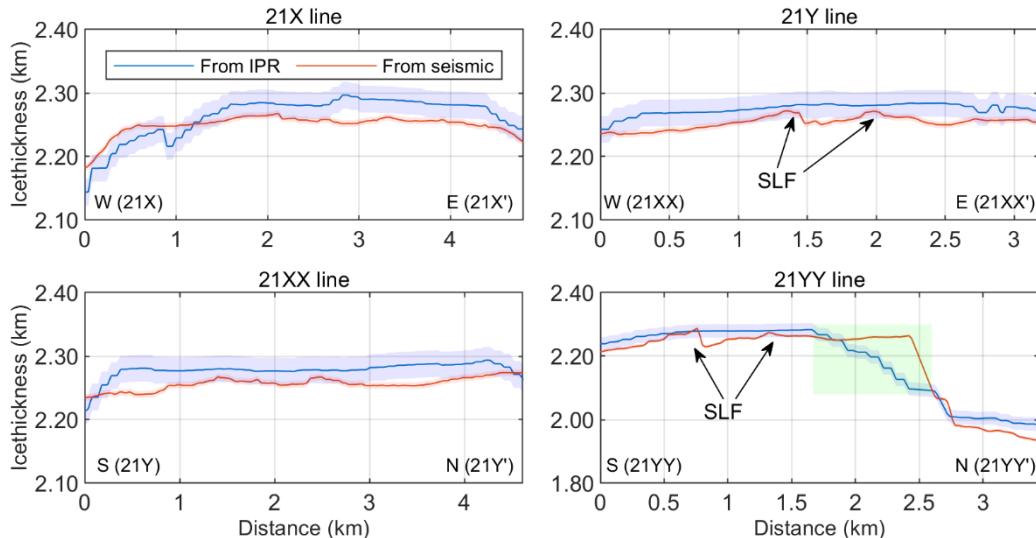
339 Figure 9c presents an enlarged comparison between the field and synthetic PSTM sections, focusing on the region of lateral
340 reflection discontinuities. In the field data, discontinuous reflections and associated low reflectivity observed at approximately
341 0.7 km and 1.2 km (TWT = 1.18 s) complicate interpretation. To simulate this feature, the velocity model incorporates a
342 concave structure at the beneath the glacier, representing the SLF. The resulting reflection patterns in the synthetic section
343 closely resemble those observed in the field data. In the field data, the reflection from the water–sediment interface (③) is
344 weak and poorly defined, resulting in significant interpretational uncertainty. This is attributed in part to the diffusive interface
345 of the unconsolidated upper sediment, which weakens reflection strength. Moreover, the short temporal separation between
346 reflection ③ and the preceding ghost reflection (②) results in significant waveform interference, often producing a single,
347 high-amplitude composite signal. Overlap between reflections ② and ③ leads to destructive interference, further
348 complicating the identification of the interface. In some areas, the ghost reflection (④) is unaffected by such interference,
349 allowing for an indirect estimation of the ③ interface using ghost travel-time differences. However, due to the difficulty in
350 clearly resolving this interface throughout the dataset, Model 2 was constructed using a uniform geometry derived from the
351 average time interval between reflections ① and ③, corresponding to a water depth of 10 m and a sediment thickness of 120
352 m. As a result, the arrival time and waveform characteristics of reflection ③ in the synthetic data exhibit slight discrepancies
353 when compared to those observed in the field data.

354 Figure 9d presents a magnified comparison of regions synthetic and field to examine reflections from a dipping bed. Within
355 2.4–2.55 km and TWTS of 1.08–1.17 s, reflections are temporally dispersed, resulting in a shadow zone where coherent signals
356 are absent. A noteworthy feature is the polarity reversal of reflections ⑦ and ⑧ between the field and synthetic datasets. In
357 the velocity models, the ⑦ interface is defined as either the glacier–bedrock interface in Model 1 or the glacier–sediment
358 interface in Model 2. In Model 1, the bedrock has significantly higher acoustic impedance than the overlying ice, due to its
359 greater density and seismic velocity, resulting in a high amplitude reflection with normal polarity. Conversely, the sediment
360 layer in Model 2 is assigned a lower seismic velocity but higher density relative to glacial ice, yielding a slightly higher
361 impedance and thus a reflection of lower amplitude. However, the field data show that reflection ⑦ exhibits reversed polarity,

362 suggesting the presence of subglacial sediments with lower acoustic impedance than assumed in the models. This discrepancy
363 may be explained by the presence of a dilatant till beneath the glacier, which can produce reverse polarity reflections depending
364 on its physical properties. Booth et al. (2012) demonstrated that the seismic response of such tills is highly sensitive to
365 variations in P-wave velocity, density, and thickness. In particular, their study showed that when the till forms a thin layer,
366 reverse polarity reflections may occur. While the existence of such glacial sediments presents a plausible interpretation for the
367 study area, the absence of reliable constraints on their seismic properties precluded their incorporation into the velocity models
368 used in this study.

369 To further validate the interpretation, ice thickness estimates from the seismic data were compared with those derived from
370 airborne IPR surveys (Ju et al., 2025) along four seismic lines (Fig. 10). Given the lack of spatial coincidence between seismic
371 and IPR profiles, kriging-based two-dimensional interpolation (Isaaks and Srivastava, 1989) was applied to the IPR dataset to
372 estimate the ice thickness at seismic line locations. The uncertainties associated with the IPR and seismic datasets are ± 20.98
373 m and ± 5.27 m, respectively, resulting in a combined uncertainty of ± 24.05 m. The root mean square error (RMSE) between
374 the two datasets is calculated as ± 29.4 m, exceeding this expected uncertainty range. This discrepancy is attributed primarily
375 to smoothing effects introduced by interpolation in the IPR data, particularly between 1.7 and 2.6 km along line 21YY, within
376 the light green shaded area in Fig. 10, where seismic data reveal a significantly steeper basal slope. When this localized region
377 is excluded, the RMSE is reduced to ± 24.8 m, approximating the combined uncertainty. Thus, apart from localized artifacts,
378 the seismic and IPR datasets exhibit strong agreement. This consistency supports the mutual reliability of both methods and
379 validates their integrated application for subglacial lake characterization. Despite localized differences, the overall ice
380 thickness estimates from both datasets are in strong agreement, and this cross-validation reinforces the robustness of the
381 seismic interpretation and affirms the consistency between the two geophysical approaches.

382 As additional supporting evidence for this interpretation, a steeply dipping (⊖) bedrock interface observed along the 21YY
383 line is consistently identified in both the seismic PSTM profile (Figure 9d) and the IPR-derived ice thickness graph (Figure
384 10), indicating a similar topographic transition in both datasets. This interface is interpreted as a structural margin delineating
385 the lateral extent of SLD2 and likely functions as a hydrological barrier. The structural congruence observed in both seismic
386 and radar data underscores the effectiveness of integrating these datasets to delineate the boundaries of subglacial lakes,
387 particularly in regions characterized by complex basal topography.



388

389 **Figure 10:** A comparison of ice thickness estimates derived from seismic and kriging-interpolated IPR data (Ju et al., 2025) along
 390 the four seismic survey lines reveals high overall consistency between the two datasets, despite localized discrepancies. The light
 391 green shaded region in the 21YY line represents areas where interpolation contributes to the divergence between the two
 392 measurement approaches. The light blue envelope represents the uncertainty bounds associated with the IPR-derived estimates,
 393 while the light red envelope indicates uncertainty bounds for the seismic-derived estimates.

394 6 Conclusion

395 Since 2016, the Korea Polar Research Institute (KOPRI) has conducted a series of geophysical investigations to study SLD2
 396 (Subglacial Lake Cheongsuk) beneath David Glacier, beginning with airborne IPR surveys. In 2021, a seismic survey was
 397 carried out to characterize the internal structure and water column of SLD2. The field seismic data revealed a strong, reverse
 398 polarity reflection at the glacier–lake interface. In contrast, the basal reflections beneath the lake are less well-defined,
 399 suggesting the presence of subglacial sediments. This ambiguity gives rise to two alternative interpretive scenarios based on
 400 the presence or absence of a sedimentary layer.

401 Given this interpretational ambiguity regarding the sediment layer, two velocity models were constructed: Model 1, which
 402 assumes the absence of sediment, and Model 2, which includes a sediment layer beneath the lake. Synthetic seismology was
 403 generated using wave propagation modeling based on these models. Sediment thickness in Model 2 was uniformly assigned
 404 using the average time difference calculated from selected areas of the dataset. Comparisons between the synthetic and field
 405 PSTM sections show consistent TWT times and polarities for key reflection events at the glacier–lake interface, the lake–
 406 bedrock interface in Model 1, and the sediment–bedrock interface in Model 2. Nevertheless, synthetic data generated by
 407 modeling a velocity model that simplifies a complex geological structure has limitations in thoroughly explaining the entire
 408 waveform of the complex field data. For example, subglacial sediments are generally expected to produce normal polarity
 409 reflections due to acoustic impedance contrasts with overlying water. However, in field data, the polarity and clarity of the

410 water-sediment interface vary with the degree of sediment consolidation. In particular, the reverse polarity reflection observed
411 at the ice–sediment interface in the 21YY profile suggests the potential presence of dilatant till.
412 This study demonstrates the utility of seismic surveying for analyzing structural characteristics of subglacial lake environments
413 that are not identified with conventional radar. Furthermore, the integrated analysis of seismic and synthetic data provides a
414 quantitative structural model of the SLD2-A geometry beneath David Glacier. These results provide critical guidance for future
415 clean hot-water drilling. In particular, we identify an area within a 1 km radius of S 75.422°, W 155.441° as a suitable candidate
416 site, based on its broad spatial extent, minimum estimated water depth exceeding approximately 10 m, and absence of
417 contamination from surface field camps. Furthermore, we plan to conduct follow-up studies incorporating advanced processing
418 techniques such as deghosting, amplitude variation with offset analysis, and the development of a refined velocity model that
419 accounts for detailed firn-layer properties. These technical advancements are expected to enhance the resolution and precision
420 of seismic imaging and contribute to a deeper understanding of the subglacial environment.

421 **Data availability**

422 The ICESat-2 data used in this study are available from the National Snow and Ice Data Center (NSIDC). The seismic data
423 and ICESat-2 laser altimetry data used in this study are also available from the Korea Polar Data Center (KPDC) upon request
424 at <https://dx.doi.org/doi:10.22663/KOPRI-KPDC-00001177>. The maps related to Antarctica were created using the
425 Quantarctica dataset version 3.2 (Matsuoka et al., 2018).

426 **Author contributions**

427 HJ: Writing—original draft, investigation, methodology, conceptualization. SGK: Writing—original draft, methodology,
428 conceptualization, supervision. YC: Writing – original draft, data processing, modeling. SP: Data processing methodology.
429 MJL: Writing – original draft. HK: Hot-water drilling. KK: Investigation. YK: Investigation. JIL: Project administration,
430 Funding acquisition.

431 **Competing interests**

432 The authors declare that they have no known competing financial interests or personal relationships that could have appeared
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437 seismic surveys. We name Subglacial Lake D2 Subglacial Lake Cheongsuk (SLC). The name Cheongsuk has a significant
438 meaning, as it is the pen name of Dr. Yeadong Kim, the founder of the KOPRI and former president of the Scientific Committee
439 on Antarctic Research (SCAR). Dr. Kim personally led the IPR and seismic surveys of Subglacial Lake Cheongsuk and
440 coauthored this paper.

441

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444

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